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# PLAINS TECTONICS ON VENUS

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Tectonic deformation in the plains of Venus is pervasive, with virtually every area of the planet showing evidence for faulting or fracturing. This deformation can be classified into three general categories, defined by the intensity and areal extent of the surface deformation: distributed deformation, concentrated deformation, and local fracture patterns. Each of these styles offers information about the tectonic history of the surface, as well as the physical properties and processes of the crust and upper mantle. Distributed deformation is manifested as individually narrow wrinkle ridges, troughs, and fractures which occur in subparallel sets that commonly extend over hundreds of km. The orientations of these sets tend to be generally constant over large regions, providing a valuable indicator of the regional stress and implying a basal (e.g., mantle convection) origin to this stress. Cross-cutting relationships with other geologic features (such as craters and volcanic landforms) and among different sets of fractures in the same region can provide relative timing information about the evolution of the surface. Concentrated deformation occurs in deformation belts, the nature and origin of which arc problematic. The characteristics of ridge belts are generally consistent with an origin related to a regional compressive stress, although some evidence suggests that their formation is not simply related to a regional stress field. It is not evident how fracture belts formed, with both extensional features and elevated topography; the process by which deformation is concentrated into belts is also unclear. Local fracture patterns range in character from faint, parallel fracture sets to polygonal fracturing reminiscent of cooling or desiccation cracking (albeit on a much larger scale). The nature. spacings, rind lengths of these features can offer insight into the mechanical properties of the uppermost layers of the crust and mechanical lithosphere. New flow lows available for dry diabase have important implications for modeling tectonic features and using them to infer the structure of the crust and mantle. In particular, these results call into question the importance, or even the existence of a weak lower crustal channel.

## I. INTRODUCTION

Plains are the most widespread geologic province on Venus, inking up over 80% of its surface. Originally, based on the relative flatness of the non-highland areas of Venus at the resolution of the Pioneer Venus altimeter, plains on Venus were defined in terms of elevation (Masursky et al. 1980), comprising all terrain below roughly the 1.5 km contour. With the higher resolution provided by Magellan, most of this area has indeed been found 10 be surfaced by flat, generally radar-smooth units believed to be duc to flood volcanism, analogous to the volcanic plains found on the Moon, Mars, and (probably) Mercury. In [his chapter wc are concerned with the tectonic deformation of these plains, particularly that deformation that is not intimately associated with highlands, tessera, or volcanic features such as shield volcanoes and coronae.

One of the surprises to emerge from the Magellan images was the ubiquity of tectonic deformation evident in the plains. Almost no area is free from some sort of fracturing or faulting. This is in sharp contrast to the other terrestrial planets. On the Earth, tectonic deformation tends to be concentrated near plate boundaries. Similarly, albeit for different reasons, there are large areas on the Moon, Mars and Mercury which do not display any recognizable deformation at all.

The existence of deformational structures in a region provides a tool, in addition to such things as craters and volcanic flows, to unravel the temporal geologic sequence of events through superposition and cross-cutting relationships. Tectonic features also provide key information on the structure of the crust and lithosphere. This is particularly important for Venus, for which we have no subsurface seismic information or very high-resolution regional gravity data As discussed later, the length scales of tectonic features, which presumably formed due to horizontal extension or contraction of lithospheric layers of varying thickness and mechanical competence, can be used in combination with geologic observations and experimental information on rock strength at varying pressure-temperature conditions to constrain the thicknesses and vertical strength distribution of the crust and lithosphere. Thus, the study of the tectonics of Venus' plains can offer important insights contributing to our understanding of the history and processes of its crust and upper mantle.

Primarily for organizational purposes wc divide tectonic features observed on the surface of Venus into three categories: distributed deformation, concentrated deformation, and local fracture patterns. These divisions are based on the intensity and areal extent of deformation. Distributed deformation consists of sets of features which individually have a small, but discernible, amount of contraction, extension or shear, and which in aggregate over an extended region can comprise a significant strain. Concentrated deformation is manifested in quasi-linear zones of intense deformation, separated by areas of relatively undisturbed terrain. Local fracture patterns are sets of features with limited areal extent, and whose individual features have widths at or below the resolution of the Magellan imaging system (no better than ~100 m). Thus this type of deformation is not expected to contribute significantly to the larger-scale strain of the crust.

## II. DISTRIBUTED DIWORMA'1"1ON

### A. Wrinkle Ridges

Wrinkle ridges are long, narrow, sinuous features ranging in width from the limit of SAR resolution to about 1 km. and with lengths up to several hundred km (Fig. 1). They generally are brighter on Magellan SAR images than the surfaces upon which they occur, and [his brightness contrast is unrelated to the orientation of the ridges with respect to radar look direction (McGill 1993), implying that the enhanced radar return is more related to wavelength-scale surface roughness than it is to topography. Most Venusian wrinkle ridges occur in sets of approximately evenly spaced, parallel ridges. It is common for two, and sometimes three, of these sets to occur in the same area. Where more than one set occurs, it generally is clear that they are of different ages, breed on stratigraphic relationships or on the intersection relat ionships among individual ridges of the several sets (McGill 1993).

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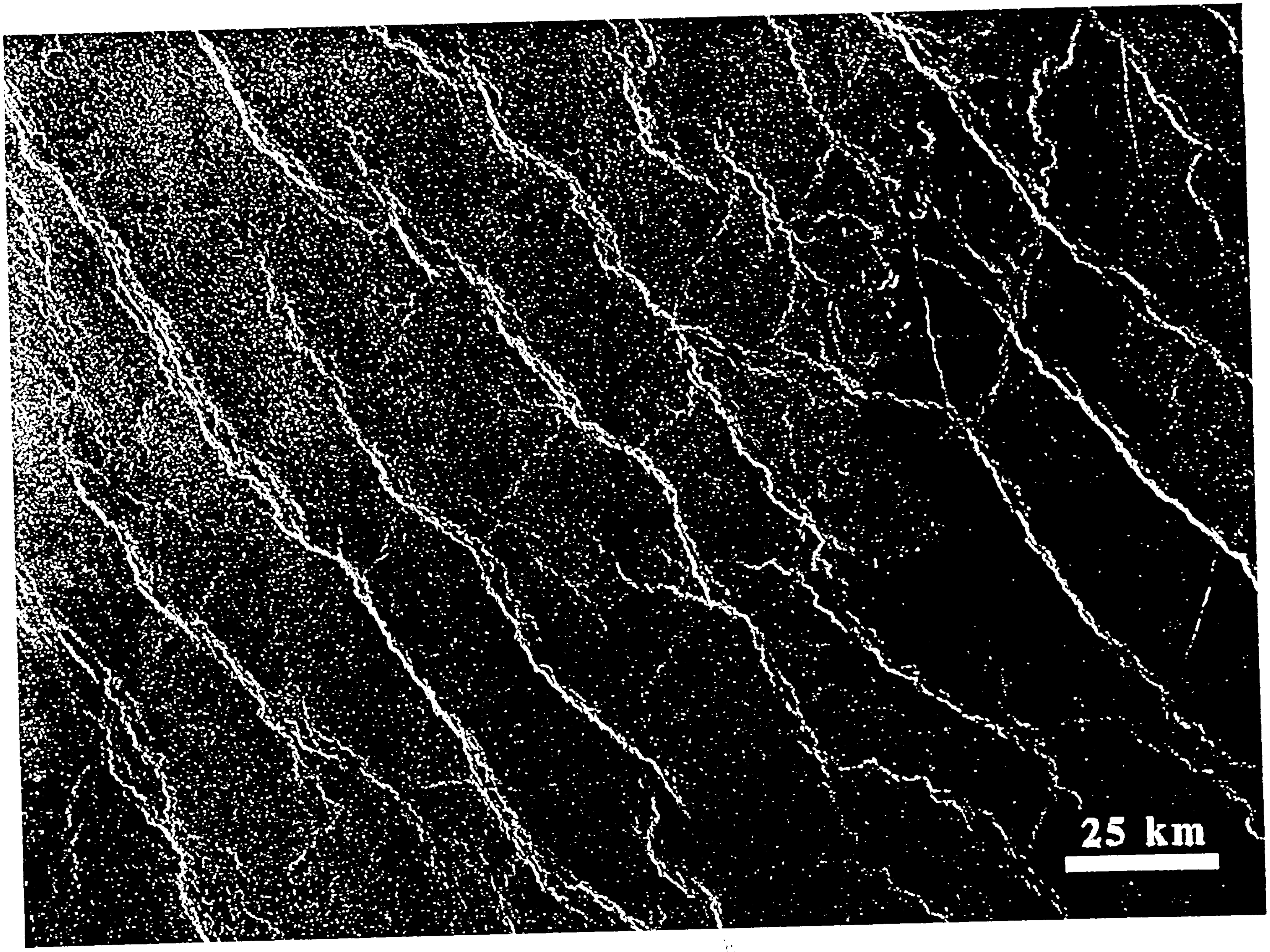
Figure 1. Typical wrinkle ridges in Rusalka Planitia (177°E, 2.5°N). Ridges are sinuous, less than 1 km wide, and exhibit an average spacing of about 20 km; part of C1-MIDR00N180, tile 19.

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*Distribution.* Almost all plains areas on Venus arc characterized by abundant wrinkle ridges. As plains constitute approxim ately 80% of the surface of Venus (Masursky et al. 1980), wrinkle ridges are extremely abundant on Venus; more so than on any other planet or moon, Wrinkle ridges are evidently not present (or not discernible) on tessera terrain, and are very rare or absent on lobate and digitate lava flows associated with relatively young shield volcanoes. They also appear to be less common in areas of intense post-plains deformation and within mountain belts.

*Identification as Wrinkle Ridges.* The primary criteria for the identification of wrinkle ridges on Venus are the long, narrow, sinuous morphology, and the occurrence in organized sets (Plescia and Golombek 1986: Watters 1991). Some of the larger wrinkle ridges exhibit the lateral brightness contrast on SAR images expected of ridges, but most features interpreted as wrinkle ridges on Venus hove insufficient relief to cause resolvable shadows or bock-slope darkening on SAR images. Some wrinkle ridges hove ponded flows derived from impact, and others appear to define sharp contacts of local plains units (see, e.g., Figs. 1 and 2 of McGill 1993). These relationships imply positive relief, and thus it is inferred that other planimetrically similar sinuous features also are positive relief features. Wrinkle ridges on other planets commonly have a complex morphology characterized by a gentle arch surmounted by a much narrower, sinuous ridge that can be in the center of the arch or to either side (Strom 1972; Bryan 1973; Maxwell et al. 1975: Watters 1988). Some Martian and lunar wrinkle ridges involve an even broader, subtle rise (Lucchitta 1977; Plescia and Golombek 1986). Venusian wrinkle ridges either do not include broad rises and gentle arches, or these features are topographically too subdued to affect the radar return. Consequently, Venusian wrinkle ridges appear comparable to the simpler forms found on Mercury, Earth. the Moon and Mars. Wrinkle ridges generally are much smaller than the ridges or ridge belts (see below), and wrinkle ridge trends do not seem to exhibit any consistent relationship to those of ridge belts on a global scale; in places they are parallel, elsewhere they are not. Local i y, however, features that appear to be wrinkle ridges grade into ridges of ridge belts by a gradual increase in width and apparent relief along trend. In addition, many coronae and corona chains include radial and concentric structures that appear to be morphologically similar to typical wrinkle ridges found on the plains. Our discussion of the structural anti tectonic significance of wrinkle ridges is confined to the well-defined sets of parallel features found on the plains.

The origin of wrinkle ridges has been under debate for at least three decades. Both igneous and structural hypotheses have been proposed, as reviewed by Plescia and Golombek ( 1986). Recent research has narrowed the controversy to two main models: (i) buckling with or without some faulting (Watters 1991); and (2) thrust or reverse faulting with or without some folding (Plescia and Golombek 1986; Golombek et al. 1991). Both of these models imply that wrinkle ridges result from compressive stresses in the crust that are oriented approximately normal to the length of the ridges.

There is some independent evidence on Venus supporting this inference where wrinkle ridges interact with well-defined topographic features (see, e.g., Fig. 3 of McGill 1993).

*Age Relations.* Almost all Venusian wrinkle ridges are superposed on the materials that make up the plains, and thus the ridges must be younger than these plains. Witbin the plains, the abundance of wrinkle ridges commonly decreases with decreasing age of specific plains units, suggesting a progressive contractional deformation of crustal rocks during emplacement of the plains materials (Solomon et al. 1992; Squyres et al. 1992). In some places, wrinkle ridges appear to postdate at least the initial stages of corona formation (McGill 1993, 1994), but this relationship has not been systematically tested globally. In contrast, young lobate and digitate lava flows, especially those associated with large shield volcanoes, are nearly devoid of wrinkle ridges. Although the relative ages of impact crater's and wrinkle ridges are commonly ambiguous, almost all ridges that can be dated relative to craters are older than the craters. Thus wrinkle ridges appear to be temporally related to plains formation, with most ridge-forming deformation occurring relatively early, based on the limited crater data available. During the waning stages of ridge formation the trends of ridge sets maintained the same orientation in some places, e.g., parts of Lavinia and Guinevere Planitiae (Solomon et al. 1992: Squyres et al. 1992), whereas in other places trends changed significantly, e.g., Eistla Regio (Basilevsky 1994; McGill 1994).

*Structural Domains.* Most plains regions of Venus are characterized by a dominant, throughgoing set of wrinkle ridges that maintains a roughly uniform trend or uniform curvature of trend over hundreds to thousands of km. For example, in most of Lavinia Planitia the dominant trend is northeast (Squyres et al. 1992), over a very large area of Sedna and Niobe Planitiae the dominant trend is approximately east-west, in Rusalka Planitia the dominant trend is northwest (Lancaster and Guest 1994), and within Aino Planitia the dominant wrinkle-ridge set defines a large arc concentric to Artemis Chasma (McGill 1992; Bilotti and Suppe 1992; Bilotti et al. 1993). These regions represent stress domains that are presumably confined to the shallow crust, Although studies are in progress (see Fig. 2 for a map of ridge directions in the Aphrodite/Eistla region), a global map of stress domains defined by wrinkle ridges has not yet been completed. In some areas, such as Aino Planitia, the dominant wrinkle-ridge set appears related too major tectonic or topographic feature (Bilotti et al. 1993). In other areas, such as Sedna and Rusalka Planitae (Lancaster and Guest 1994), the dominant trend is clearly oblique to nearby tectonic and topographic features. In many, and perhaps most, plains areas there is a second (red, rarely, a third) set of wrinkle ridges in addition to the dominant set that defines the domain. In all areas so far studied in sufficient detail, these additional sets are younger than the dominant set. In many places, the younger sets are clearly related to local tectonic and topographic features. For example, on the plains adjacent to Eistla Regio younger wrinkle ridges define sets concentric to the large shield volcanoes (Basilevsky 1994; McGill

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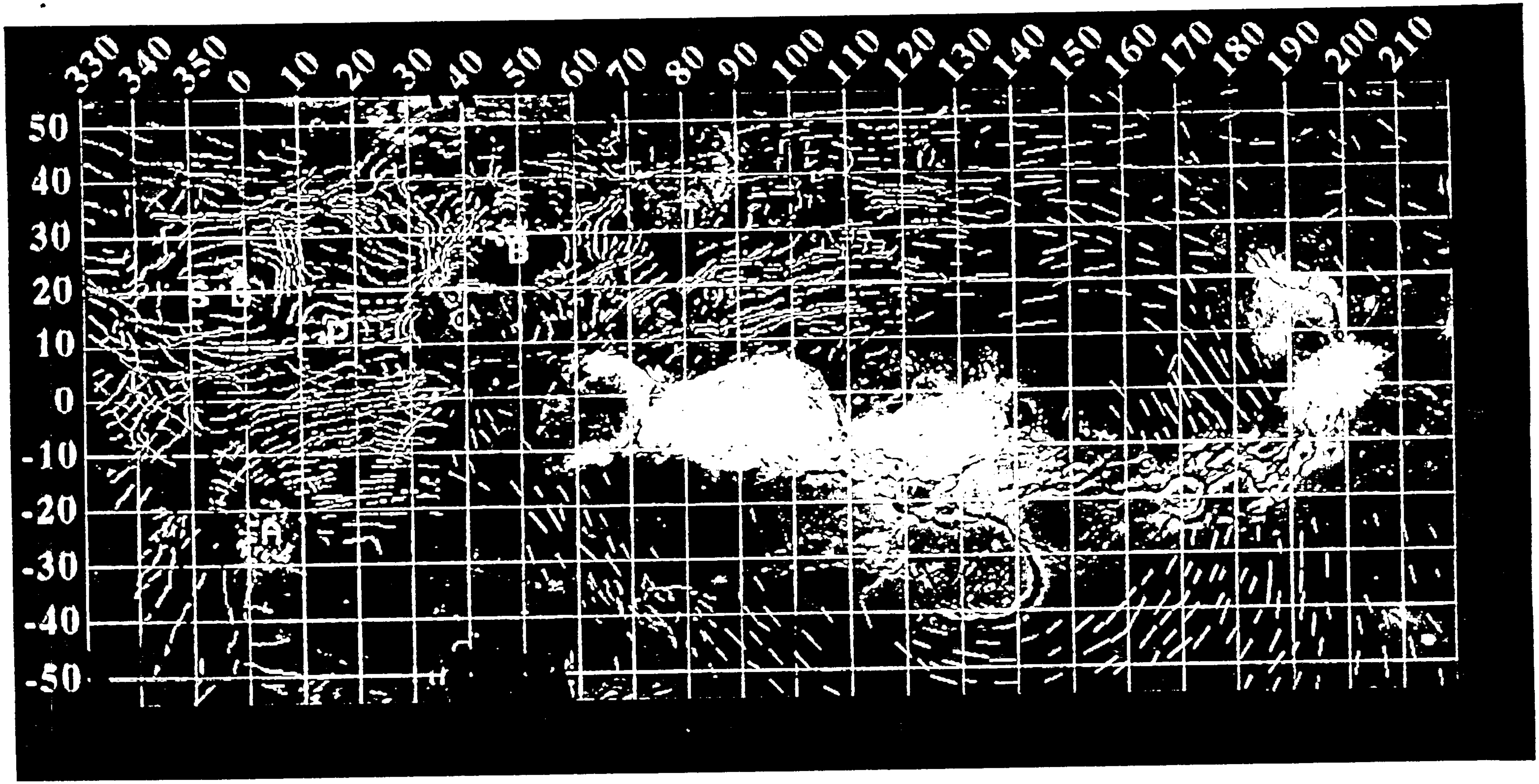


Figure 2

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1994; Copp and Guest 1995).

The terns of km spacings typical of many wrinkle ridges are indicative of a depth penetration of regional compressive stresses on the order of the thickness of the strong upper crustal layer (i.e., a few to 10 km). The observed widths of 1 km or less, which are generally maintained along strike for tens to hundreds of km, imply small amounts (~1- 5%) of upper crustal shortening.

### B. Grabens, Linears, and Dikes

Linear features of extensional or inferred extensional origin are abundant on the Venusian plains. Many of these show the brightening or darkening on SAR images that one would expect for slopes facing towards or away from the radar; these are inferred to be fault scarps. Many of these scarps occur in facing pairs bounding troughs that are morphologically identical to grabens on Earth, Mars and the Moon. More common are radar-bright linears that are too narrow to define their boundaries because they are only one or two pixels across. This limited pixel width also means that any relief associated with these line.ars is not resolvable.

Very few areas underlain by the global plains are completely devoid of faults or linears. Plains Surrounding volcanic constructs and coronae are especially likely to have abundant extensional structures, commonly with multiple tl-ends. Large (width greater than two km or so) grabens seem to be more common adjacent to tessera terrain.

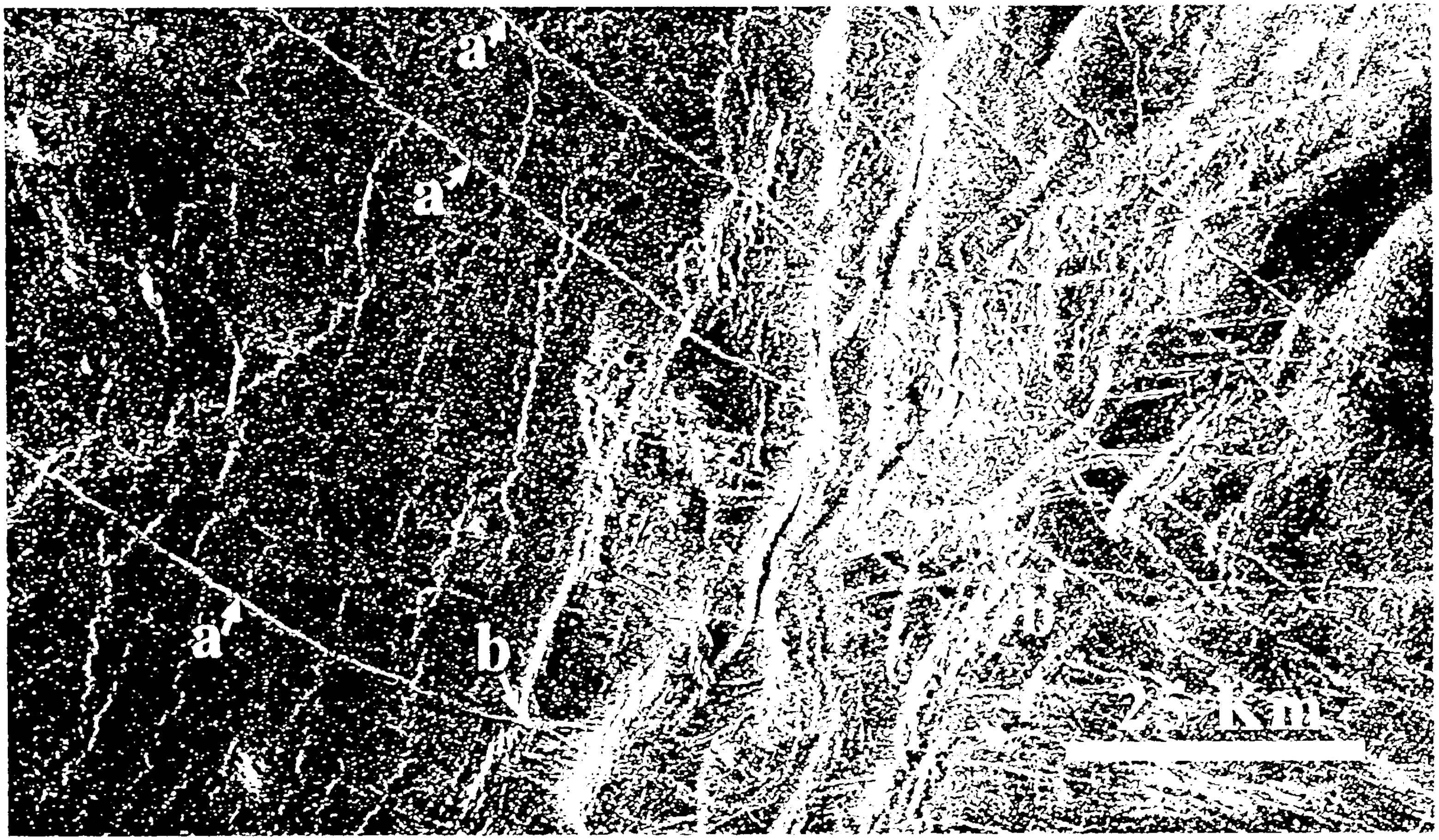
Scarps, especially those bounding troughs, are readily interpreted as extensional by direct analogy with similar features on Earth. The narrow line.ars are inferred to be extensional as well, based on less direct evidence. Because many of these features parallel larger structures resolvable as grabens or fault scarps, and because some increase in width along trend into scarps or grabens (Fig. 3), it is very likely that all of them represent extensional structures. In addition, the planimetric patterns defined by some of these linears suggest origins by extension, as will be discussed in more detail below. Thus they could be either fractures or small faults. Because the brightness of these features on SAR images seems unrelated to their orientation with respect to the radar look direction, this brightness is most likely due to enhanced roughness at radar wavelength scale rather than due to topography. This characteristic would seem 10 favor interpreting narrow linears as fractures (joint zones?) rather than as fauns.

Many extensional structures occur as radial or concentric sets associated with coronae or volcanic centers. Individual linears of these sets extend hundreds or even thousands of km from the source feature. These radial sets very likely represent surface fractures or narrow grabens overlying dikes derived from the central feature, analogous to the well-s[udied examples in northwestern Scotland and the Spanish Peaks region of Colorado (Anderson 1951; Odé 1957; Muller and Pollard 1977). McKenzie et al. (1992 ) present a mechanical analysis of the Venusian radial and concentric linears based on models developed for these terrestrial examples. At large distances from the central source, figure to be reduced & pasted up @ P.b. house. Fig 3 follows

Figure 3. Examples of typical bright, narrow linears with unresolvable geometry (points "a") that widen along trend into resolvable grabens (points "b"). Thus these linears, and parallel members of the same set, are interpreted as due to extensional strain; parts of FMIDR45S350, tiles 9 and 10.

the radial sets commonly follow preferred trends that most likely define the regional trajectory of the maximum principal compression. These sets locally diverge around holes in the crust (McKenzie et al. 1992), a pattern consistent with the behavior of compressive stress trajectories around holes in elastic plates. In the vicinity of Aphrodite Terra, the inferred maximum compression is normal to the long wavelength topography (Grosfils and Head 1994), in general agreement with the results of Bilotti et al. (1993) based on trends of wrinkle ridges in Aino Planitia. Elsewhere, however, there is no globally consistent relationship between the inferred stress orientations and large-scale topography (Grosfils and Head 1994), a result that also is consistent with the wrinkle ridge data. Where available, stratigraphic relationships indicate that linears of these large radiating swarms formed after the regional plains but before almost all impact craters (Grosfils and Head 1995). Assuming that the determinable relative ages are characteristic of the global relationships, this implies that the dikes inferred lie at depth beneath the linears, and thus also the source centers, all formed very soon after emplacement of the global plains.

In addition to the large swarms of linears associated with well-defined central sources, the plains regions of Venus commonly are cut by much shorter linear features. These can occur as closely spaced, parallel linears generally scores to hundreds of km long, or as en-echelon lines a few km to a few tens of km long. Most commonly, there is no connection between these smaller

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linears and topographic, structural, or volcanic features. Locally, these smaller linears are so closely spaced that they define a fabric that is penetrative at the scale of the Magellan images (see, e.g., Squyres et al. 1992, Figs. 3 and 14). It seems unlikely that these smaller linears can be explained as the surface manifestations of dike emplacement. Some of these small linears are parallel to sets of larger extensional structures radial to coronae or volcanic centers. and thus they probably are simply mode I (purely extensional) cracks formed in response to the same regional stress field that is controlling the far-field orientations of the radial swarms. Other small linears are not parallel to larger structures, but their similarity in size and geometry to sets that are parallel to larger structures suggests that they, too, represent mode I cracks.

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Most of the grabens occurring as members of linear sets are narrow, ranging in width from the limit of resolution to perhaps 2 km. Grabens with widths up to 15 km are common in tessera inliers within plains. Many of these are partially filled by plains materials, indicating that they are older than plains formation. However, there are local indications of continued motion on these structures after emplacement of plains (see, e.g., Fig. 33 of Solomon et al. 1992), suggesting that the stress regime responsible for these grabens still existed at the very beginning of plains formation.

### C. Broad-Scale Vertical Deformation

Vertical deformation of the lithosphere can result in characteristic tectonic patterns. These patterns are formed by a combination of extensional and contractional structures that either follow the vertical deformation gradient or are orthogonal to it (e. g., radial or circumferential orientation for a circularly symmetric deformation), depending on the horizontal scale of the uplift or depression and the thickness of the elastic lithosphere (see Banerdt et al. 1992). Such patterns have been tentatively identified around a number of relatively small features, such as volcanic edifices and coronae. On a broader scale (say, comensurate with highland plateaus) these types of pattern have not been identified, and either they do not exist on Venus or the tectonic complexity obscures them. However, the existence of lava channels and digitate lava flows provide us with snapshots of local slope directions at the time of their emplacement that can be compared with present slopes to infer vertical deformation.

Both lava channels and digitate lava flows can provide direct evidence of paleoslopes if the initial flow direction is unambiguous. Even if the flow direction is ambiguous, these features can serve as reference surfaces because the direction of slope cannot have reversed during emplacement; any slope reversals now present must be due to deformation of the original slope at a scale smaller than the entire flow or channel length. Lava channels prove especially useful in this regard because many are very long and thus define an initially gentle slope. Published results suggest that many long lava channels have experienced significant post-emplacement deformation with as much as 2 km of relief between adjacent low and high points along the channels (parker et al. 1992; Komatsu and Baker 1994: McLeod and Phillips 1994). The relief defines two scales of deformation: one with a wavelength of a few thousand km, the other with a wavelength of a few hundred km. The large-scale undulations have been defined for only the two longest channels (Komatsu and Baker 1994); these authors infer that this deformation scale corresponds to that of large basins. The shorter scale corresponds closely with the characteristic spacing of ridge belts (Zuber 1986: Frank and Head 1990; Squyres et al. 1992).

Lava channels must, of course, be younger than the plains units they transect. Although a complete global evaluation of their ages has yet to be completed, channels commonly arc disrupted by impact craters and cut by wrinkle ridges (McGill 1993; Komatsu and Baker 1994). This suggests that they formed either late in the global episode of plains formation, or very soon thereafter. Thus channels appear to be relatively old plains features.

## 111. CONCENTRATED DEFORMATION

Several of the plains regions of Venus are characterized by long, narrow belts of relatively intense deformation. The dimensions of these belts vary widely; widths range from narrow tips a few km across to broad zones as much as 300 km wide. and lengths range from < 100 to 2000 km or more (Frank and Head 1990; Senske et al. 1991). Most of these belts stand a few hundred meters to more than a km above the surrounding plains. although a very few deformation belts occur within shallow depressions. Deformation belts tend to be bright on SAR images compared to adjacent plains, and this contrast is probably duc both to topographic effects and to the greater roughness of the belts at radar wavelength scale. Initial studies of these features were based on Venera 15 and 16 images (Barsukov et al. 1986; Basilevsky et al. 1986; Kryuchkov 1988; Sukhanov and Pronin 1989; Sukhanov et al. 1989; Frank and Head 1990) and Arecibo images (Campbell et al. 1991; Senske et al. 199 1). Because of the very low incidence angle of the Venera radar, and the kilometer-scolc resolutions of both data sets, it was difficult to determine if the individual linear structures within these belts arc scarps, graben-like grooves, or ridges (see, e.g., Senske et al. 1991). The better resolution and more favorable incidence angles of Magellan images allow a clear distinction in most instances between deformation belts characterized by scarps and grabens, referred to as "fracture belts," and deformation belts characterized by ridges, referred to as "ridge belts" (Solomon et al. 1992; Squyres et al. 1992). Although it is convenient and logical to discuss fracture belts and ridge belts separately, as we do, there are composite belts in places.

### A, Ridge Belts

*Description and Classificatio n.* Ridge belts arc morphologically diverse. Some consist essentially of a single broad arch, with smaller ridges superposed in some places. More commonly, ridge belts include a complex array of individual ridges that also exhibit diverse morphologies. The most thorough descriptions of ridge belts, and the only classifications, are based on Venera 15 and 16 images (Kryuchkov 1988; Frank and Head 1990). Kryuchkov (1988) divided ridge belts into three broad classes: (1) belts consisting essentially of a single wide swell with a broad summit (Fig. 4); (11) belts made up of many closely spaced smaller ridges (Fig. 5); and (III) "spaced" ridges, with individual ridges very far apart compared to their widths. Classes I and II are easily recognizable on Magellan images, but it is not clear that there is a distinct group of belts corresponding to elms 111 (see below).

Frank and Head (1990) developed a classification of individual ridges based on Kryuchkov's three classes of ridge belts. Class I ridge belts are termed "broad arches," and consist of single arches 20 to 40 km wide, commonly with smaller' ridges superposed. Based on Venera images, these ridges were considered 10 be analogs of lunar, Mercurian and Martian wrinkle ridges (Frank and Head 1990), which commonly consist of a broad arch with a narrower superposed sinuous ridge. This analogy is less convincing on Magellan images because there generally are several superposed smaller ridges rather than one, and because these superposed smaller ridges are less sinuous and significantly broader than typical Iirst-older wrinkle ridges on other planets (Watters 1988). It is clear on the Magellan images of Nephele Dorsa (Fig. 4), a class I ridge belt or broad arch (Frank and Herd 1990), that the superposed small ridges most similar 10 lirs[-order wrinkle ridges arc, in fact, members of a wrinkle-ridge set on the adjacent plains [hat cross the ridge belt at a low oblique angle. The superposed small ridges confined to Nephele Dorsa are larger and straighter than typical wrinkle ridges.

Class II ridge belts were subdivided into three subclasses by Frank and Head (1990) according to whether the individual ridges within the belts are discontinuous, paired, or anastomosing. Mos[ class 11 belts are dominated by anastomosing patterns of individual ridges, but discontinuous and paired ridges are generally interspersed among the anastomosing ridges. The overwhelming majority of all ridge belts are in class 11. These also are the largest ridge belts, with widths as great as 300 km and lengths up to several thousand km 10 COI1Y. Individual ridges within class 11 belts arc typically 5 to 15 km wide, with a slightly larger inter-ridge spacing (Frank and Head 1990; Squyres et al. 1992). In Vinmara and Atalanta Planitiae class II ridge belts are sufficiently abundant to determine the typical spacing between them. Spacing varies from 70 to 670 km, but most values lie between 325 and 425 km (Zuber 1986; see also Fig. 4b of Frank and Head 1990). The ridge belts of Lavinia Planitia exhibit similar spacings of ridges and bells (Squyres et al. 1992).

Frank and Head (1990) did not provide a detailed description of a class 111 ridge belt. The illustrated example is the same one used by Kryuchkov (1988), incorrectly located in Bezlea Dorsa. The area is actually in easternmost Sedna Planitia, about 25° to the west of Bezlea Dorsa. The ridges shown are widely spaced, sinuous, and < 1 km wide: they are part of a completely typical set of figure to be reduced & pasted up @ pub. close. Fig follows.

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Figure 4. Northern part of Nephele Dorsa, a Class 1 ridge belt located at 40°N, 140°E in northern Niobe Planitia. Nephele is dominated by a broad, gentle arch about 40 km wide. The narrow, sinuous ridges superposed on this arch in the southern half of the figure are wrinkle ridges associated with the surrounding plains, and are not related to Nephele. The. ridges superposed on the brighter part of the arch in the northern half of the figure do appear to be related to Nephele. However, these are wider than typical wrinkle ridges and not sinuous; C1-MIDR45N138, tiles 37 and 45.

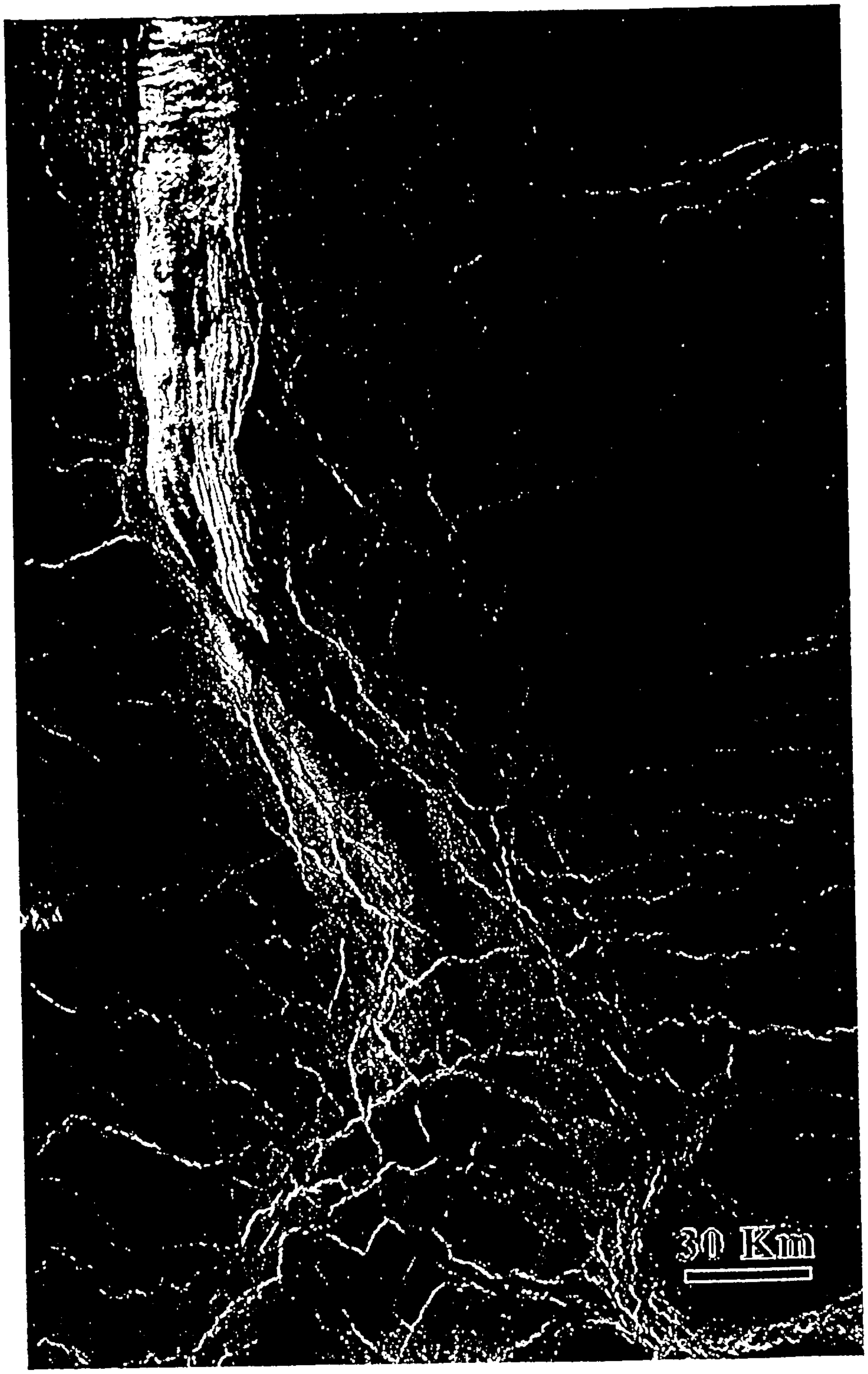


Figure 4

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plains wrinkle ridges. Thus [here is reason to doubt if this is a valid class of ridge belts.

*Age and Origin.* The most thoroughly studied areas of ridge belts are in Atalanta, Vinmara, and Lavinia Planitiae. In Lavinia Planitia, the ridge belts have deformed a radar-bright, "textured" plains formation that clearly is older than the materials characteristic of most of the global plains (Squyres et al. 1992, Figs. 3 and 14). The ridges and the textured plains are embayed by flood-type lavas. It thus seems clear that in this area the ridges formed relatively early in [he evolution of the plains. In the northern hemisphere, however, relationships arc more complex. In Vinmara Planitia, the relative ages of ridge-belt and plains materials are ambiguous for many, and perhaps most, of the class II ridge belts (Rosenberg 1995). Some ridge belts in the Atalanta-Vinmara area deform tessera terrain, as do belts south and cast of Fortuna Tessera (Sukhanov and Pronin 1989), but these belts also extend across and are locally younger than adjacent plains. As discussed briefly in Solomon et al. (1992), it thus is not clear whether (a) ridge belts always formed early in plains development but plains in different areas are not the same age, or (b) ridge belts form-xl at diverse times in the development of coeval plains.

Ridge belts arc generally interpreted as resulting from global-scale compressive stresses oriented normal to their [rends (Basilevsky and Head 1988; Basilevsky et al. 1986). Tbc arch-like morphology and topography of some individual ridges (Barsukov et al. 1986; Frank and Head 1990) supports the interpretation of these features as anticlinal folds. The common tendency for ridge belts to be elevated relative to their surroundings is also consistent with localized crustal thickening due to the same globol-scale stress field (Solomon et al. 1992; Squyres et al. 1992). The elevated Topography will induce gravity sliding stresses (Turcotte and Schubert 19 S2.), and the importance of these stresses can be estimated from the predicted strength of the lithosphere and the elevation of the ridge belts over the surrounding plains. This can be as much as 2 km (Ford and Pettengill 1992; Frank and Head 1990), but is more typically hundreds of meters (Squyres et al. 1992). As discussed below, the depth-averaged strength of the Venus lithosphere can be assumed to be controlled by Byerlee's Law in the brittle regime and by flow laws for dry diabase (Mackwell et al. 1995) for the crust and dry olivine (Chen and Morgan 1990) for the mantle in the ductile regime. The maximum elevation of the ridge belts results in a maximum gravity sliding stress of ~50 MPa, and, for the average strength indicated by laboratory experiments (see below), the stresses could produce failure in the upper several km of the Venus lithosphere. Gravity sliding stresses associated with the elevation of individual ridge belts may thus contribute to the development of shallow, sm:ill-scale deformation features within the belts. But given [he greater strength of the Venus lithosphere at depth (which would inhibit longer wavelength deformation due to gravity sliding) and the fact that ridge belts occur in areas of average or low elevation relative to the global mean, broader-sc:dc ridge belt deformation requires figure to be reduced & pasted up @ pub. dept. Fig follows

Figure 5. Pandrosos Dorsa, a ridge belt in Vinmara Planitia. (a) Northern, relatively simple portion of the belt (214°E, 63.5" N). The belt here is 40 km wide; the individual ridges are about 1.5 km wide and spaced about 5 km apart; part of C1-MIDR60N208, tile 14. (b) Southern, complex portion of [he belt (207.5°E, 56°N). The width of the belt is 220 km; individual ridges are similar to, but more closely spaced than ridges in the northern portion. Abundant bright linears (fractures?) are present in addition to ridges, and the belt exhibits evidence for a prolonged history of continuous or alternating deposition and deformation; parts of C1-MIDR60N208, tiles 44 and 45.

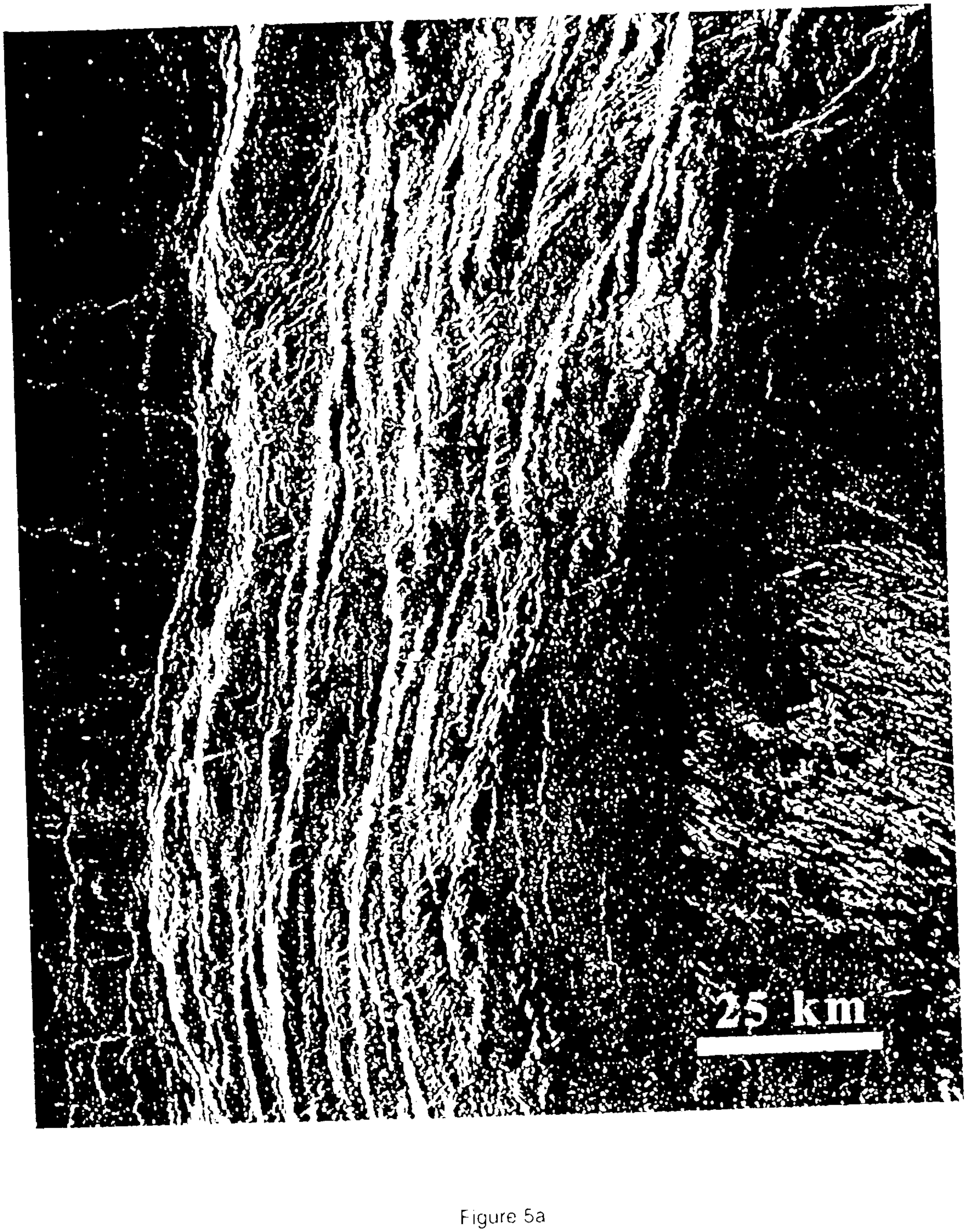


Figure 5a

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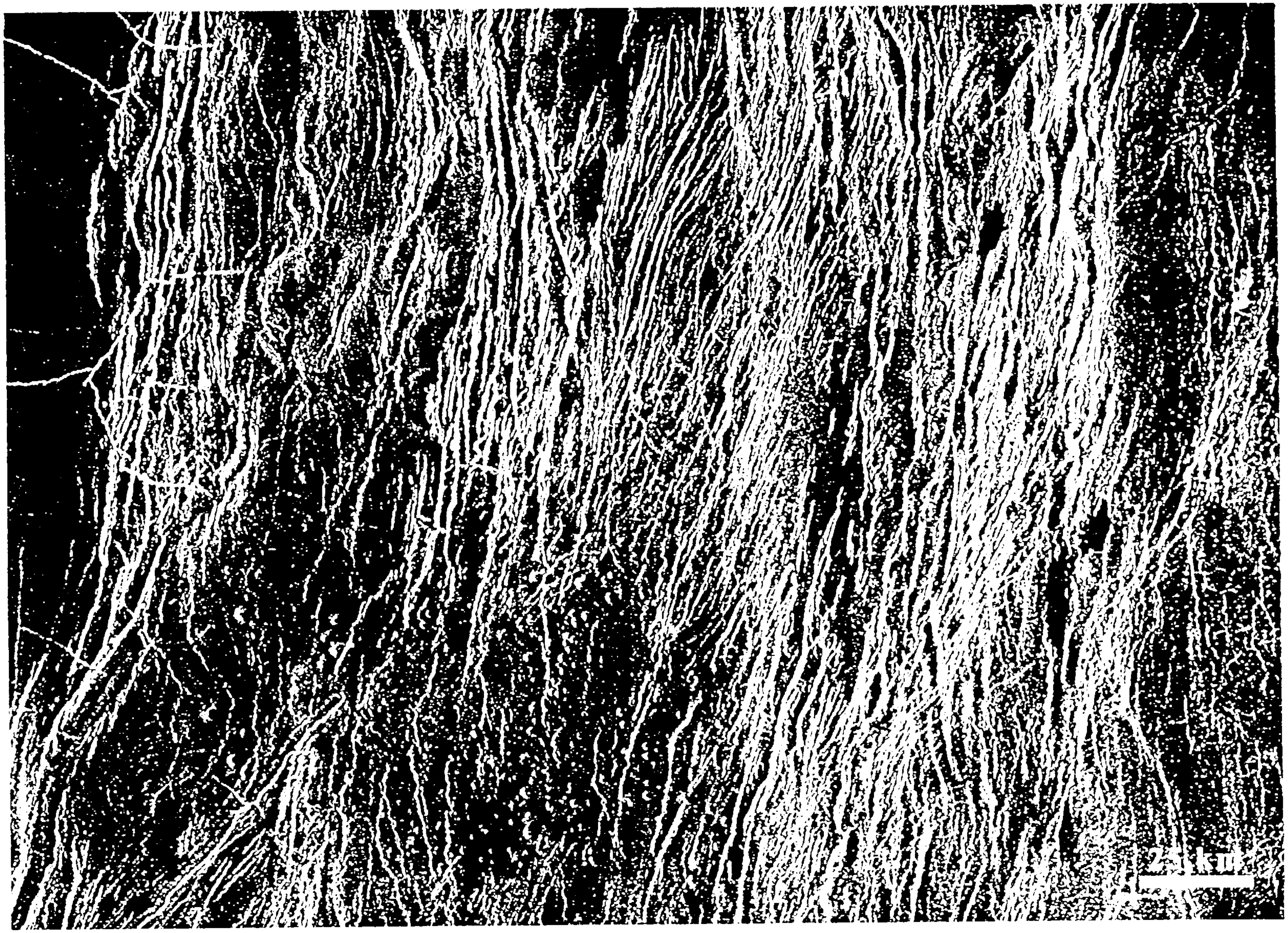

Figure 5b

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an additional or alternative mechanism, such as mantle flow-related stresses (Phillips 1990). Given the broad, spatially coherent style of deformation of some ridge belts (Zuber 1990), the source of stress that produced these features is consistent with formation by mantle downwelling (Phillips et al. 199 1; Squyres et al. 1992: Zuber 1990). The lack of a broad, regional uplift due to thickening of the crust that would be a consequence of the downwelling process (Bindschadler et al. 1992) indicates that the ridge belt fan in Atalanta Planitia could be a site of incipient downflow (Zuber 1990). It has also been proposed that ridge belts may mark the sites of former downwelling, with belts formed due to thrusting associated with rebound subsequent to the cessation of downward flow (Phillips et al. 1991).

An alternative explanation of at least some ridge belts as due to extension has been proposed (Kryuchkov 1990; Raitala and Tormanen 1990: Sukhanov and Pronin 1989) based on the recognition) that features such as volcanic centers associated with some ridge belts are difficult to explain in a compressional stress environment. However, the defining characteristics of these belts as ridge-like arches would not be expected in a tensional stress regime. Tensional bending stresses at the crests of flexural folds produced due to remote compression may provide an explanation for at least some of the observed localized extension. A flexural (as opposed to penetrative deformation) response of the lithosphere to compressive stress would be consistent with the notion of significant mechanical competence of the lithosphere, as suggested from gravity modeling (see, e.g., the chapter by Phillips et al.).

## B. Fracture Belts

*Description.* Fracture belts are broad, elongate swells or arches that are characterized by numerous roughly belt-pmllcl linears, scarps, and grabens. In detail, the fractures and faults of these belts commonly occur as sets that intersect at low angles. The most thoroughly studied examples occur in Guinevere and Lavinia Planitiae in the southern hemisphere (Solomon et al. 1991, 1992: Squyres et al. 1992). Fracture belts in Guinevere and Lavinia Planitiae range in size up to about 200 km wide and 1000 km long, and stand a few hundred meters to more than a km above adjacent plains. Globally, it is difficult to clearly define fracture belts, because in places structures similar to those described in Lavinia Planitia grade along trend into corona chains (Fig. 6) or ridge belts (see, e.g., Figs. 29 and 30 of Solomon et al. 1992).

*Age and Origin.* Where mapped in Lavinia and Guinevere Planitiae, fracture belts clearly fault and elevate materials of the surrounding plains (Squyres et al. 1992; Solomon et al. 1992, Figs. 31 and 32). Because these plains embay ridge belts, the fracture belts of Lavinia Planitia are clearly younger than the ridge belts (Fig. 7). However, young digitate flows in Lavinia Planitia were diverted by pre-existing fracture belts and thus, like ridge belts, fracture belts appear to be older than the youngest flows on Venus.

The linears (fractures'?), scarps, and grabens making up fracture belts are all consistent with formation by tension normal to the belt trends. The

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Figure 6. Fracture belt in northern Bereghinya Planitia, located at 50°N, 20°E. This belt, which is morphologically similar to those found in Lavinia Planitia, grades to a chain of coronae to the south; C2-MIDR30N026, tiles 4, 5, 12, and 13.

presence of rhomboidal depressions bounded by faults on some fracture belts (Solomon et al. 1991) suggests that many are actually due to transtension. On the other hand, the broad ridges characteristic of these belts suggests crustal thickening due to compression, as for ridge belts (Squyres et al. 1992). But the ridge and fracture belts of Lavinia Planitia are nearly orthogonal to each

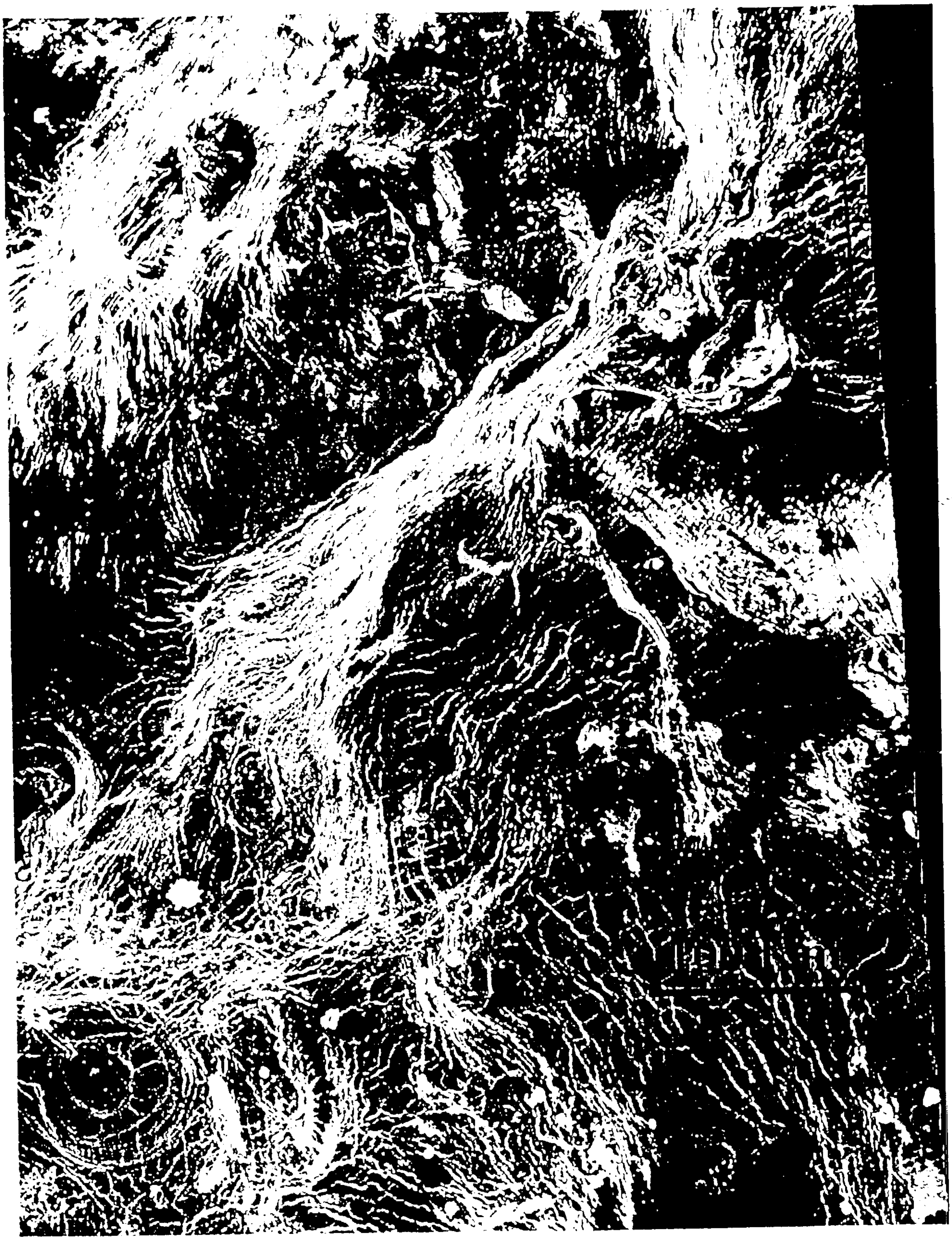

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Figure 7. Ridge and fracture belts in Lavinia Planitia (348°E, 38°S). The ridge belt, trending NE, is 30 km wide and generally consists of 2 or 3 individual ridges 3.5 to 4 km wide. The ridges and the material [hey deform appear to be embayed by the surrounding plains material. The transecting fracture belt, trending NW, cuts the ridges and the surrounding plains, and [bus is younger than the ridge belt. Note that the wrinkle ridges on the plains arc generally not parallel to the ridges of the ridge belt: C1-MIDR45S350, tile 4.

other (e.g., Fig. 7). It is possible in Lavinia Planitia, at least, to account for this apparent paradox by assuming a nearly 90° reorientation of principal compression trajectories between formation of ridge belts and fracture belts. However, this appears inconsistent with the persistence of NNE trends of wrinkle ridges. If fracture belts are due to compression rather then tension, then the characteristic extensional structures must result from local tension due to uplift and bending (Solomon et al. 1991). However, for a typical belt width of 200 km and height of 1 km, simple sinusoidal bending would have induced a maximum strain of only about 1% (for a 40 km plate), insufficient to account for the abundant fractures and faults.

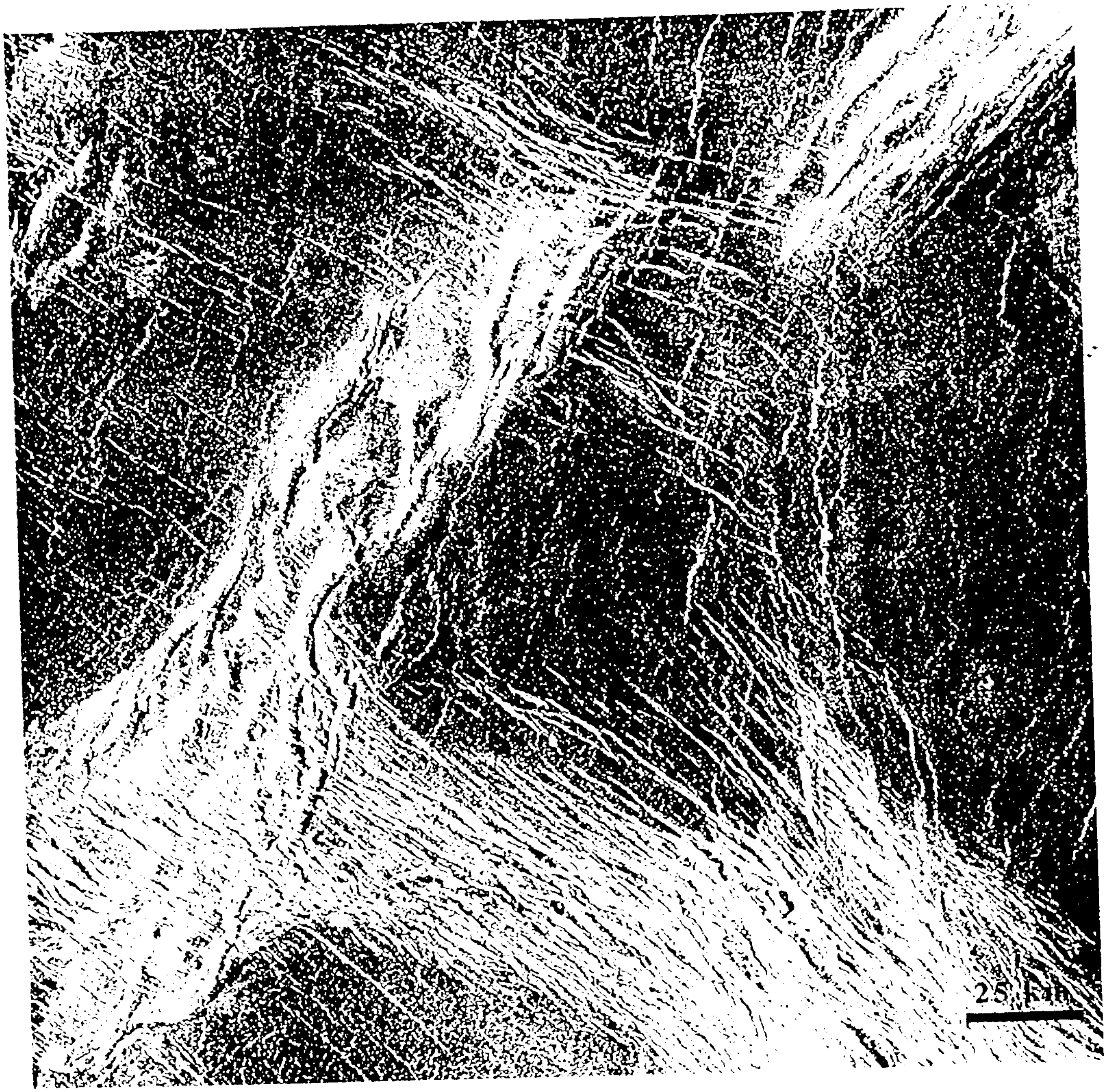


Figure 7

## IV. LOCAL DEFORMATION

Local deformation as defined above is commonly manifested on the Venusian plains in patterns of line-scale structures, which are developed on the kilometer to sub-kilometer scale. These patterns can be classified generally into parallel sets, polygonal patterns and irregular patterns.

### A. Parallel Fracture Sets

Solomon et al. (1991) first noted remarkably linear features which arc developed with a regular spacing of about a km in the "gridded plains" of Guinevere Planitia, where faint, regular lineations form the NE-trending component of the grid (scc Fig. 3 of Solomon et al. 1991 and Fig. 1 of Banerdt and Sammis 1992). Although such a well-developed orthogonal grid is unique to this location, Banerdt and Sammis (1992) found similar sets of regularly spaced lineations in many locations on Venus. and concluded that such features arc relatively common throughout the plains.

An example of such a set is shown in Fig. 8. They arc composed of parallel, thin (a single pixel in Magellan images), straight lineations whose microwave reflectivity dots not depend on radar illumination direction. The patterns typically cover mess with dimensions of hundreds of km. The average spacing of lineations is small, between 1 and 2.5 km, and the scatter in individual spacings is about ±1/3 the average. Based on these observations, especially the narrow, linear geometry and the azimuthal independence of radar reflectivity, Banerdt and Sammis (1992) concluded that they are tension fractures in the brittle upper layers of the volcanic plains material.

The very close spacing of these features is perplexing, as conventional geophysical models for regular spacing require an unreasonably thin lithospheric layer (< 1 km; Solomon et al. 199 1). Onc possible solution to this problem is suggested by the observation from structural geology that jointing within a sedimentary layer often occurs with a spacing roughly proportional to the layer thickness (see, e.g., Pollard and Aydin 1988). Calculations of the state of stress around a vertical crack (see, e.g., Lachenbruch 1961; Pollard and Segall 1987) show that the relief of horizontal tensile stress (the "stress shadow") occurs mostly within a distance comparable to the crack depth, inhibiting the development of subsequent cracks within that region. However, Banerdt and Sammis ( 1992) observed that these patterns appeared to have virtually the same spacing (1–2.5 km) everywhere they were observed. They proposed a shear-lag model in which a relatively thin sur-face layer is partially decoupled from similar material below by a frictional contact. This results in a spacing between features that is independent of the thickness of the layer, as both the frictional resistance and the layer strength scale similarly with thickness. An implicit requirement of this model is that the layer have a relatively large tensile strength (implying only a small amount of pre-existing fracturing). This implication and the observation that the parallel fracture sets do not appear to follow other superimposed structural trends suggest that the

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Figure 8. Typical parallel fracture pattern on plains units in Eistla Regio (15° N, 44°E); parts of FMIDR15N043, tiles 29,30,37, and 38.

tectonic episode responsible for the parallel fracture set must predate other strong deformation events recorded on that surface. Thus these structures may serve as a relative temporal marker for a deformation sequence, showing the orientation and sense (extensional) of the earliest tectonic event to which the plains unit was subjected.

### B. Irregular Structures of the Gridded Plains

The set of features orthogonal to the thin parallel lineations in the gridded plains have been -the subject of several studies which have attempted to use their morphology and length distribution to constrain lithospheric properties (Sammis and Banerdt 1991; Banerdt and Sammis 1992; Bowman and Sammis 1995). These features are primarily extensional in origin, because they grade into recognizable grabens to the north. However, their distinct curvilinear en-echelon morphology suggests a component of shear as well. Bowman and Sammis (1995) have inferred that these structures formed by the propagation

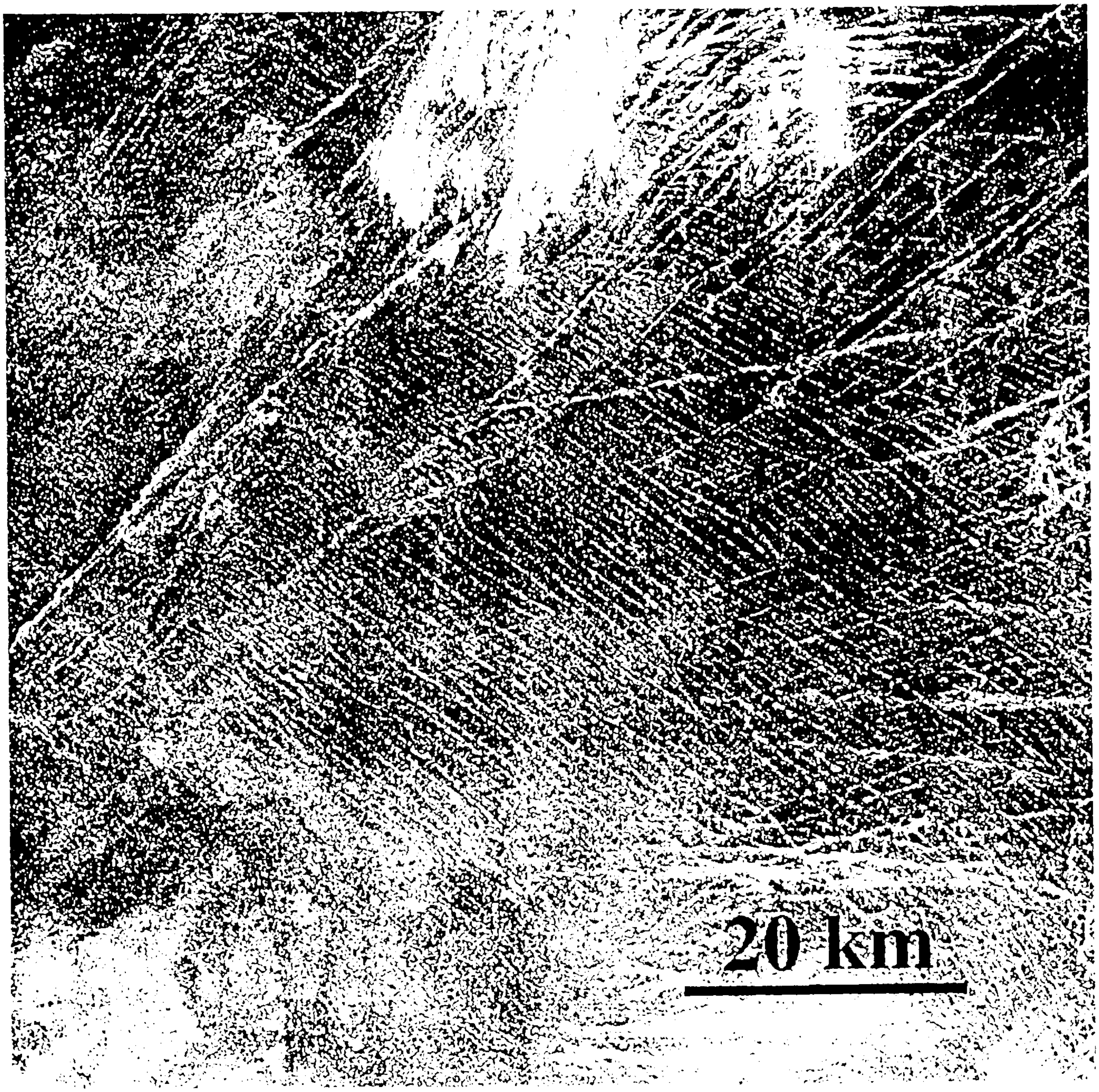


Figure 8

of deeper ("basement") fractures through a surficial basalt layer to the surface. The length distribution of these structures is found to have a distinct break in slope at a length of about 80 km. In the model of Bowman and Sammis (1995), this break corresponds to the change from the two-dimensional growth of a semi-circular crack in an half-spoce to simple one-dimensional horizontal elongation after it has penetrated to the bottom of the brittle lithosphere. With this model the length distribution implies a thickness for the mechanical lithosphere of about 40 km, in agreement with independent determinations in other areas.

### C. Polygonal Patterns

Polygonal patterns of bright lineations, broadly similar in appearance [o cooling and dessication crack patterns (albeit at a much larger scale), are another common feature on the plains (Fig. 9). These appear at scales from 10 km down to the resolution of the Magellan radar, and are typically 1 to 2 km across. Because of their isotropic orientations and apparent tensional nature, Johnson and Sandwell (1992) interpreted these features [o be due to thermal stresses. They investigated two scenarios for the generation of these stresses, cooling of an initially liquid lava flow and local heating of the lithosphere from below. Both process were found to generate sufficiently large stresses to produce the fracturing for reasonable parameters, but the reheating model was favored duc to the difficulty in scaling the dimensions of the polygons from the meter-scale structures seen in terrestrial lava lakes to the kilometer-scale features observed on Venus.

There are also common occurrences of more complex and irregular patterns of radar-bright lineations with length scales of the order of a few km (Solomon et al. 1991; Johnson and Sandwell 1992). In some of these areas a background grid of parallel lineations is discernible (Banerdt and Sammis 1992). It is possible that processes similar to those responsible for the kilometer-scale spacing of the parallel fracture patterns also control the length scales of the more irregular patterns.

## V. DISCUSSION

The likely lack of a low viscosity zone in the Venus mantle (Kiefer et al. 1986) implies that mantle convective stresses could be strongly coupled to the overlying lithosphere. Numerical experiments have demonstrated that mantle flow-related stresses transmitted to the lithosphere could attain magnitudes sufficient for tectonic deformation (Phillips 1990). It is thus prudent to compare the distribution of plains tectonism to the internal density structure implied from gravity and topography data. Global gravity fields (Reasenberg and Goldberg 1992; McNamee et al. 1993; Nerem et al. 1993; Konopliv et al. 1993; Konopliv and Sjogren 1994) and a regional high-resolution line-of-sight inversion (Barriot and Balmino 1994) show the plains of Venus to be relatively gravitationally featureless. This is not surprising, as tbc resolution of

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Figure 9. Polygonal patterns in Guinevere Planitia (21°N, 334°E). The lineations are locally obscured by presumably younger volcanic domes and associated deposits (e.g., "a"), and are truncated at the boundary between lighter and darker plains units ("b"). Brighter, more continuous lineations trending NE (e.g., "c") are parallel to, and in places grade into, wrinkle ridges in the adjoining plains. We interpret them to be younger than the polygonal pattern, as they appear to follow the polygonal grid; parts of FMIDR20N334, tiles 20,21,28, and 29.

the gravity is much less than the length scales of individual tectonic features. Topographically, the plains are, by definition, at or below the planetary mean (Ford and Pettengill 1992; Rappaport and Plaut 1994), which is generally consistent with the.ubiquity of contractional tectonic structures.

Global models of the internal density distribution from inversion of gravity and topography data (Banerdt 1986; Her-rick and Phillips 1992) show evidence for isolated upwellings among an interconnected network of downwellings. Major regions of downwelling are all associated with plains units, some of which contain ridge belts. Power spectral ratios of gravity and topography are also consistent with a large-scale potter-n of downwelling flow

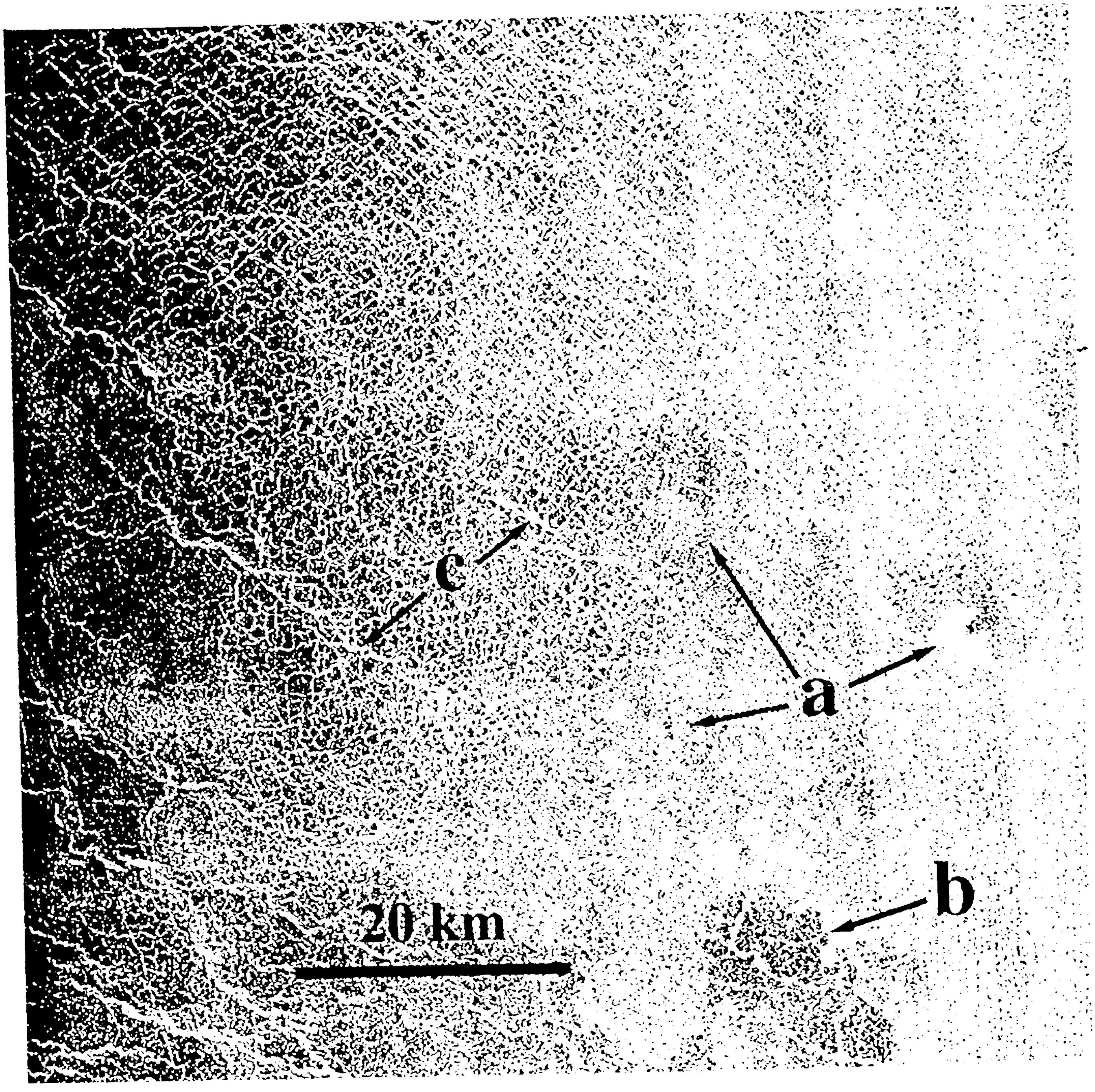


Figure 9

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beneath [he plains lithosphere. An alternative interpretation is that the plains are regions of crustal {binning associated with mantle uplift (Buck 1992), but this hypothesis has not been tested in the context of the observed distribution of surface tectonics. Gravity/[ opography relations of plains regions indicate a thickness of the mechanical lithosphere of ~50 to 100 km (Bindschadler 1994; Simons et al. 1994; Banerdt et al. 1994), which must be reconciled with the presence of both long and short length scales of tectonic features in the plains.

The pre-Magellan view that Venus exhibits a thin mechanical lithosphere was based on observations of the length scales of tectonic features defined by stretching and shortening instabilities (Zuber 1987; Zuber and Parmentier 1990) or characteristic elastic wavelengths (Banerdt and Golombek 1988), and on the depths of impact craters as compared to models of viscous relaxation of surface relief (Grimm and Solomon 1988). These models were all characterized by common assumptions concerning the composition of the Venus crust and mantle: (1) the Venus mantle is similar in composition to Earth's mantle and therefore the primary constituent is olivine, and (2) the crust in areas where observed impact and tectonic structures arc located is similar to that determined for the Soviet Venera and Vega landers, and is appropriately described by [he mechanical properties of diabase (Surkov et al. 1983,1984,1987). Results for both classes of models indicated a range of Venus crustal thicknesses of ~10 to 30 km and associated thermal gradients of <25 K km⁻¹. However, the results are critically dependent on knowledge of the brittle and ductile deformational behavior of diabase and olivine.

Numerous experiments on terrestrial rocks indicate that the brittle strength of near-surface rocks is essentially independent of rock type. strain rate and grain size (Byerlee 1968); strength depends almost solely on pressure (depth) and is described in a simple linear relation by Byerlee's Law. The ductile strength of crustal and mantle materials is significantly more problematic, as it is sensitive to temperature, strain rate, composition, and modal mineralogy (Kohlstedt 1992). Because of its importance with regard to flow in the Earth's mantle, the ductile rheology of single crystal olivine is relatively well understood (Goetze 1978; Kirby and Kronenberg 1987: Chen and Morgan 1990; Kohlstedt et al. 1995), albeit for strain rates many orders of magnitude greater than characterize the mantle. Crustal rheologies are much less well characterized. A particularly important issue is that the data used in pre-Magellan studies of crustal rheology were derived from experiments in which the samples were not completely dried (Shelton 1981 ; Shelton and Tullis 1981: Caristan 1982). Water remaining in the samples during deformation led to low (by terrestrial comparison) flow strengths at the nmr-surface temperature of Venus, with the depth of transition from brittle deformation by shear failure to ductile flow by ternperaturc-controlled dislocation creep occurring within a few km of the surface (Fig.10a).

Given the lack of water on Venus, at least near the surface (Kaula 1990), experiments performed at exceptionally dry conditions arc applicable. Recent delete title

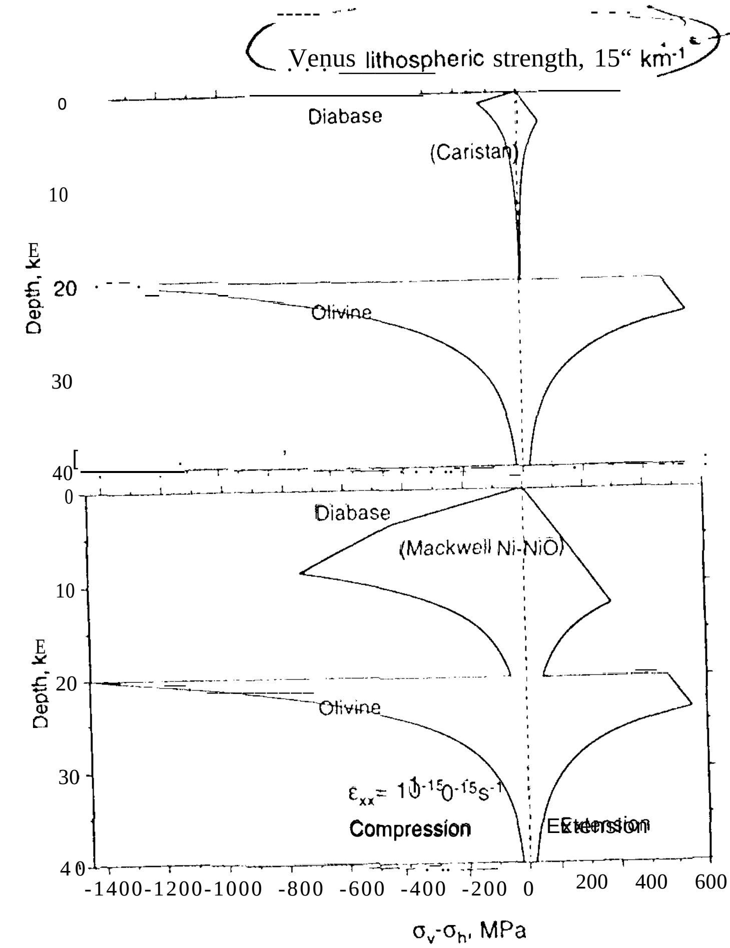


Figure 10. Strength profiles of& Venus lithosphere in unifonrr horizontal compression and extension assuming diabase crustal compositions using flow laws from (a) Caristan (1982) and (b) Mackwell et al. (1995). Both calculations assume a surface thermal gradient of 15 K km$^{-1}$, a crustal thickness of 20 km and an olivine mantle with the flow law of Chen and Morgan (1990).

experiments have been performed on thoroughly dried samples of Columbia diabase (Mackwell et al. 1995; see also the chapter by Phillips et al.), with this rock type chosen on the basis of gross similarity to chemical compositions determined at the Venera lander sites (Surkov et al. 1987). The yield envelope illustrated in Fig. 10b indicates that the absence of water results in a Venus crust that is much stiffer than previously thought. The much higher strength is consistent with Magellan findings from topography and gravity data of a thick present-day elastic lithosphere (Johnson and Sandwell 1992,1994; Moore et al. 1992; Sandwell and Schubert 1992*a*,*b*; Phillips 1994; Schubert et al. 1994; Banerdt et al. 1994; chapter by Phillips et al.). However, one should not discount the possibility that some of the structures we now see were formed under conditions that were significantly different (such as higher crustal temperature gradients and water concentrations) than those at present.

With the new experimental data on rock rheology, a systematic re-analysis to understand the structure of the crust from the length scales of plains structures is necessary. The first step in any such analysis is the identification and quantification of widths and periodic length scales, where they exist. from Magellan SAR imagery and altimetry. This is particularly important for areas that were either not imaged before Magellan or that contain features with short tectonic length scales that were imaged at lower resolution.

Future analyses should necessarily incorporate advances in modeling techniques for the development of tectonic features. For example, previous models that related wrinkle ridge and graben spacings [0 the thicknesses of lithospheric layers utilized continuum folding (for contraction) and necking (for extension) instability models that did not take into account the almost inevitable presence of faults associated with those features. Current numerical techniques have the ability to incorporate the presence of faulting, either by *a priori* inclusion (Melosh and Williams 1989) or by strain localization (Scholz 1990) techniques (see, e.g., Neumann and Zuber 1995). Such approaches are also relevant to models of highland deformation features such as rifts and mountain belts.

A major consequence of the new experimental rheological data is the absence or at least a significant decrease in size of a weak lower crustal channel. The crustal channel was thought to act as a decoupling zone between the strong upper crust and the upper mantle and was believed to enable the simultaneous development of multiple tectonic length scales within a given region (Zuber 1987; Banerdt and Golombek 1988). Certain classes of ridge and fracture belts have well-dciined widths in association with an apparent regular development of kilometer-scale deformation. If the regularity of the small-scale fracturing can be established, the question arises how to develop multiple length scales of deformation in [he absence of a lower crustal channel that separates strong lithospheric layers near the surface and at depth. Possible explanations include the existence of other compositional or rheological layers in the crust (such as individual lava 11ow units; Banerdt and Sammis 1992), strain weakening during deformation in areas of finite strain (Zuber 1994), locally high near-surface thermal gradients, and additional mechanisms for the development of tectonic length scales. Alternatively. the length scales (especially the longer scales) may reflect period icities in the forces that formed them (perhaps related to convective processes) rather than the mechanical properties of the elastic lithosphere itself. Establishing whether long and short length scales of deformation developed concurrently would provide an important constraint on the origin of the features. While there is no current evidence contrary to the assertion that all length scales developed contemporaneously, such relationships are difficult to document from SAR images.

Another important scale observation is the persistence of tectonic trends over large areas, sometimes for thousands of km. This appears to require either a very strong mechanical lithosphere capable of transmitting stresses over large distances without undergoing large non-elastic strains, or else the processes that are causing the deformation over wide plains regions are dominated by tractions on the bottom of the lithosphere rather than edge forces.

Parallel fracture patterns on the Venusian plains contain various enigmatic elements. Interpreted by Banerdt and Sammis (1992) as tension fractures, these features have been explained using a shear-lrrg mechanism, in which the spacing of the lineations is controlled by a relation between the tensile strength of the brittle layer and a shear traction on its base. In this scenario there is no relationship between fracture spacing and the thickness of the brittle layer. Though shown to be plausible, models that incorporate layer thickness dependencies have yet to be tested. In such models it will be necessary to investigate whether the controlling layer is the brittle crust or other sub-layering. A possible alternative model also involves shear at the base of a surface brittle layer, but has the spacing of surface features dependent on the thickness and mechanical properties of the layer as well as the shear traction at the base of the layer. In either model the source of regional tension must be identified. Flexural uplift, lithospheric cooling and mantle flow have been suggested (Banerdt and Sammis 1992), but these possibilities have not been quantitatively tested.

## VI. SYNTHESIS AND FUTURE DIRECTIONS

The full value of the information available from the tectonics of Venus' plains awaits the completion of detailed geologic studies of the various plains regions, as well as the highlands, bemuse the processes which formed onc have undoubtedly affected the other. This will allow the undertaking of the global syntheses necessary to put the bewildering array of tectonic features into a consistent framework, and allow the integrated tectonic history of the planet to be inferred.

In assessing the many unique aspects of plains deformation on Venus, the effect of the absence of water on deformational style provides a natural focus for future work. Experiments relevant to the ductile strength of the Venus lithosphere. both crust and mantle, should be performed for a broader range of modal mineralogy (for the crust) and grain size, as well as for larger strains. Theoretical models should incorporate general distributions of lithospheric strength, and also must adapt to deal with combined continuum and fault deformation, as well as time dependencies. The greatest challenge in future studies of plains tectonism will be to understand how to relate the complex time history of deformation to the post-resurfacing global stress state of Venus. Such analyses will be essential to understand the complex thermal evolution of Venus and its differences from Earth.

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